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Peculiarities in the Formation of Downstream Lowland Riverbeds

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N. Y. Makkaveyev

PECULIARITIES OF THE FORMATION OF DOWNSTREAM LOWLAND RIVERBEDS

(In the articles of N. Y. Makkaveyev: "Peculiarities of the Formation of Downstream Lowland River Beds" and of D. A. Kozlovskiy: "River Bed Processes and Recent Vertical Shifts of the Earth's Core," the authors hold contradictory viewpoints on certain premises on the formation of river beds. The editorial board is considering the publication of both works,which are devoted to the vital and, generally, very controversial questions of river bed dynamics, which are of indubitable interest, in one compendium)

BACKGROUND OF THE PROBLEM

On the basic tenets of Davis's (1899) theory on the cycles of river erosion is the principle fixing the dependence of the processes of accumulation and erosion in riverbeds upon the relative changes of the erosion level. According to this principle, an inevitable accumulation ensues in riverbed bottoms upon elevation of the erosion level for a considerable stretch upstream starting from the newly formed estuary, whereas in level-lowering, deep erosion ensues and the river bottom cuts into the alluvium or the bed rock, likewise for a considerable stretch upstream. This principle is based upon the premise that the receiving basin always exerts tributary action on that part of the river stream contiguous to the estuary. Thus the level increase results in shifting the tributary zone upstream and furthers the slowing of

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the current and the accumulation of alluvium which, in turn, furthers river channel elevation.

The elevated sectors of the river bottom result, then, in a tributary of the superstratum of the river's sector, the bottom of which also gradually rises.

Intensive alluvial accumulation at the confluence of the stream with the basin and the gradual upstream shift of the accumulation zone is actually observed in rivers the discharge of which diminish toward the estuary (many rivers of arid areas). Regressive accumulation causing a gradual heightening of bottom level and water level in the river in proportion to silting of reservoirs (Shamshov, 1939) is clearly observable above many dams. Thus there are graphic confirmations of Davis's thesis and it is widely used in geometric analysis as completely accurate. However, a number of phenomena observed on rivers, particularly in lowlands, cast doubt upon the universality of this thesis.

Early in the 90's of the last century, the gifted engineer and wonderfully penetrating researcher, N. S. Lelyavskiy (1895), showed the concept of the river tributary at the point of entrance into the sea to be not wholly true because with the discharge of water which forms the bed, i.e., in spring floods, no tributary is observed in the lowlands, but rather a lowering in open surface. In his report to the Third Session of Russian Waterway Specialists in 1895, N. S. Lelyavskiy stated that with a river's issue into a basin, in proportion to a rise in river water level, an additional slope is formed which furthers the increase of flow-rate and erosion of the bottom. Thus the stream, at low water levels, when

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passing along the overly deepened bed, has low current flows and a negligible lowering of water-surface grade, i.e., the tributary phenomenon is observed only in a stream of known mean water level which does not form a river bed.

Thus on the lower Dnepr, between Kherson and Girlam the drop at mean water level is 0.32 meters whereas in the spring flood it is 2.19 meters, i.e., 7 times greater. In Lyelyavskiy's opinion the main cause was precisely that, within the limits of the indicated sector, reaches of great depths predominate.

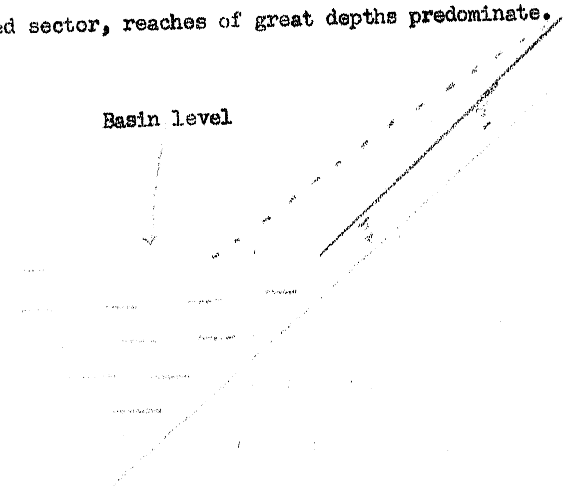

Figure 1. Tributary curve at the estuary region according to Davis h_1 actual depth h_2 depth attributary zone

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M. P. Rudskiy (1905) expressed this very idea in more general form having ascertained that the deepening of the bed "near the estuary in alluvial lowlands" is characteristic of many rivers.

N. I. Nikolayev and B. V. Polyakov (1937) stated the thesis that the origin of estuary depressions in the Volga's bed were due to the peculiarities of a hydrological condition of the estuary sector for which an increase of water surface slope was characteristic in spring flood periods. These authors also noted that the overdeepening of estuaries was a usual phenomenon and that it was observed for rivers both at settling and elevating sectors of the littorals.

K. K. Markov (1948) after detailed analysis of processes forming a longitudinal profile of a river, drew the conclusion that there cannot be a well-defined connection between changes of altitude of erosion level and the erosive activity of the river.

HYDRAULICS OF ESTUARY CONFLUENCES

The causes leading to the formation of overdeepened estuaries are easily ascertained in the analysis of the hydraulics of stream confluences with water-receiving basins. It is true that the general theory of estuary confluences still has not been worked out and, in the main, experimental data must be relied upon.

In order to show this hydraulic scheme by which Davis's hypothesis can be clarified we shall present a stream, to simplify the problem, which has a uniform bottom slope extending, without break, to the bottom of the receiving basin. (Figure 1). Omitting, in

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the first approximation, losses in the formation of thermal energy, for the extent of the estuary sector of the river, under given conditions, the following relation between depth and current swiftness can be presented:

$$\frac{v_1^2}{2g} + h_1 = \frac{v_2^2}{2g} + h_2,$$

with v_1 and v_2 -- current velocity in sections with corresponding depths h_1 and h_2, and g-- acceleration of the force of gravity. Supposing velocity v_2 in the second (estuary) section to approximate sero, then:

$$\frac{v_1^2}{2g} + h_1 = h_2,$$

whereupon h_2 exceeds h_1 for the value of velocity pressure. The resulting level increase must extend upstream according to the curve (dotted line in Figure 1), asymptotically approximating that stream surface which must be in its uniform (actual) state.

Value of velocity pressure ($\triangle h = h_2 - h_1$) is computed according to the equation $\triangle h = \frac{xv^2}{2g}$, with x-Koriolus' factor approximately equaling 1.1. In lowland rivers where median current velocity generally does not exceed 1 to 1.5 meters per second.

h is the value equal to several centimeters. In mountain streams where current velocity can exceed 5 meters per second, value $\triangle h$ can reach 1 to 2 meters. Because a loss of energy results from stream broadening, which we omit in equation (1)

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the value of level change is $\Delta h(1 + \zeta)$, with the coeffic-ient of loss, whose value is set at 0 up to --1 by various researchers.

Even with ζ approximating zero, value Δh for lowland rivers is, nonetheless, an insignificant part of the active depth of the stream (h); but because this latter factor decreases in the widen-ing of a stream, due to the spread of layers of water along the basin area, then at the area of issue into the basin there should be not an increase but a decrease in the level. The local lowering of level on the curve of drop is passed up to the upper reaches of the river where, as a result, an increase of current velocity occurs. This is a well-known phenomenon to hydro-technicians in flumes it is possible to obtain in the upper narrowed sector, some lowering of levels, after having chosen an adequately large angle of widen-ing.

Figure 2 gives the results of tests conducted by us in 1949. In a flume 26 centimeters wide and 180 centimeters deep, with a hori-zontal bottom, two longitudinal glass partitions were set up to con-stitute a rectilinear "bed" 10 centimeters wide. Each partition con-sisted of two parts, the lower halves of which (along the current) could easily be withdrawn without stopping the circulation of water. In the space between the partitions and walls of the flume, ob-lique partitions were set up making an agitation center of the lower half of the "bed". These partitions were likewise easily removed without stopping the water circulation in the flume and obtaining a sudden widening. Thus it became possible to determine very ac-curately the influence exerted by changing the "bed" width on water level and current velocity in the upper narrowed sector. In the

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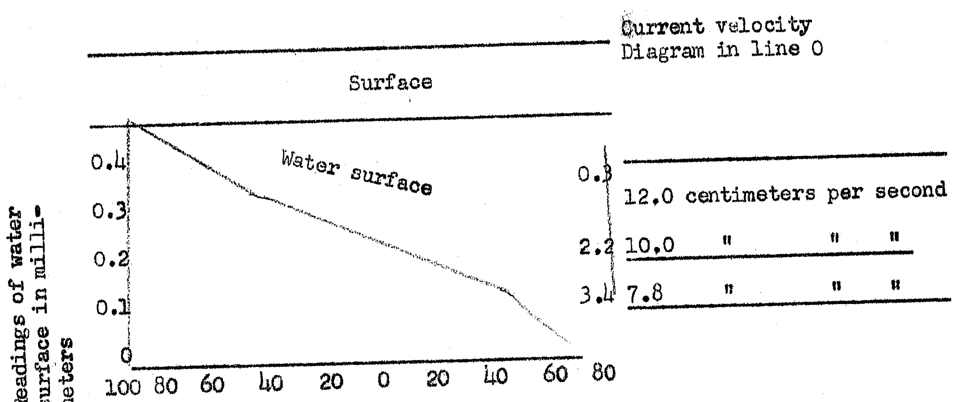
tests particular attention was paid to having the water level in the pressure reservoir absolutely constant. Measurement of current velocity was done with a tube constructed by A. I. Losiyevskiy;

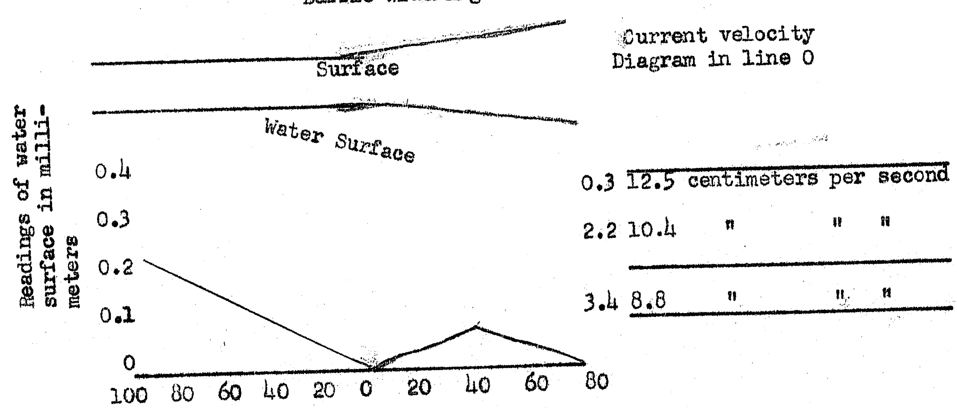
Test No 5a

Bed Walls are Parallel



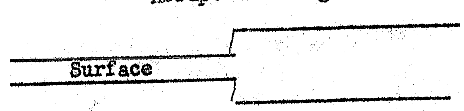
Test No 5b

Baffle Widening



Test No 5 c

Abrupt Widening



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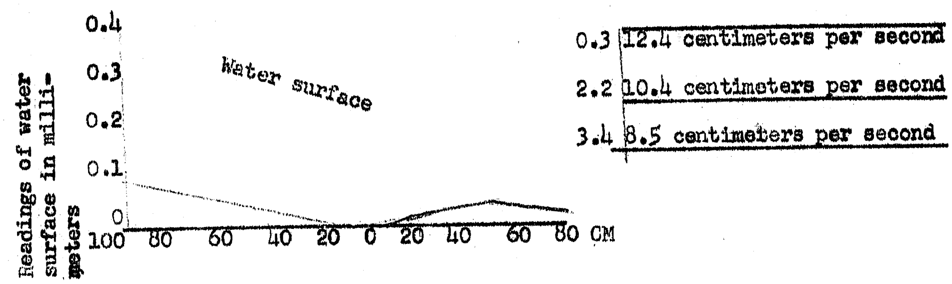

Figure 2. The lowering of water level and change in values of current velocity in widenin the stream in the rocker. Graduations on the levels have been made from an ad hoc level constant for the entire series of tests

leveling of water line was effected by a method also devised by A. I. Losiyevskiy, which enabled precision graduation up to 0.01 millimeters. The tests were repeated several times at various depths and current velocity in inverse direction, and all variants provided analogous results. Always in "bed" widening a depression of "estuary" water surface was observed, as was the lowering of water surface and the slight increase of bottom current velocity in the narrowed parts above the "estuary." In each case nothing similar to a tributary phenomenon was revealed above the "estuary."

The phenomenon of an estuary drop is particularly pointed when the bottom slope is increased below the estuary (the ledge has a steeper pitch than the trough). The phenomenon of an "estuary cascade" with the formation of a clear rise of water surface is then frequently observed (Figure 3). A similiar curved drop is observed at confluences of canals with large water reservoirs and above deep

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dredged cuts in rivers (Chertousov, 1934, page 65).

The waters of a receiving basin can constitute a stream tributary when:

(a) Current velocities of a stream are so great that Δh exceeds h_1 (mountain streams with a turbulent current); the length of the tributary zone, as P. S. Makeyev (1941) quite correctly noted, is very insignificant.

(b) The receiving basin is so small that the shore opposite the estuary stirs the tributary wave which overlaps the estuary (an artificial water reservoir);

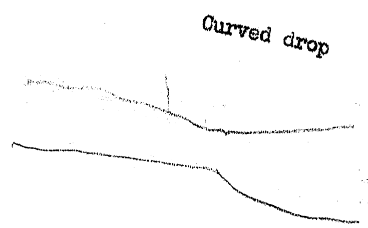

Figure 3. Curbed drop of a free surface of a stream in confluence with a canal with a deep reservoir.

(c) An independent current is in the receiving basin which prevents the spilling of the stream of the descending river (the issue of a tributary into a river having greater current velocity than the tributary).

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In all other cases the issue of a stream into a basin is attended by the phenomenon of a drop, whereupon as shall be seen later, rivers usually deepen estuary beds and form a sand bank which gradually shifts into the water area of the receiving basin, and does not rise upstream as would follow Davis's theory.

ESTUARY FORMATION

We shall now turn to the influence of the hydraulics of estuary sectors on their morphology, starting with the simplest case; the artificial canal along which a stream passes.

In hydrological engineering quite frequently (to avoid dredging) canals are joined with water-receiving basins so that the canal bottom is, at its estuary sector, level with the basin water level. When water is released along this canal, a clearly defined curved drop is formed in its lower parts and current velocities towards the estuary increase, reaching maximum values just below the estuary cut. Within the basin the velocity begins to decrease rapidly in the degree that the stream waters mingle with the still water of the water intake. The more the canal is filled, the more the curved drop broadens the distance against the current and the higher the velocity value in the estuary, where sometimes a rapid with bubbling whirlpools and high waves is formed.

We had occasion to observe a picture of a large "waterfall" at the confluence with the sea of a large canal on the Black Sea littoral with a water discharge exceeding 200 cubic meters. In the first weeks after the canal was placed in operation, current velocities in its estuary exceeded 5 meters for a second, and in the

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bed of the canal, during the passage of even small floods, erosions were observed which extended the depth of separate sectors by several meters in a 24-hour period.

The estuary waterfall, in canals with an eroded bottom, rapidly levels down because, for one, along the entire drop zone a number of deep depressions forms which gradually fuse into one or several stretches and pools, and the phenomenon of a drop becomes diffused.

Also, an accumulating fill (sand bank) forms in front of the estuary within the basin which isolates the estuary depths from those of the water intake. With low canal water levels the current, supported by the sand bank, becomes completely tranquil. The drop phenomenon lasts only in high water periods and tapers off from year to year to the degree of depth increase in the estuary zone and sand bank enlargement.

As an example of an estuary under spring flood, which estuary is still in the stage of forming a sand bank, we shall cite data on distribution of depths, slopes and current velocities along the navigable channel in the lowlands of a river where a breach of a sea littoral dune belt occurred in 1840. The breach led to the formation of a new bed along a direct course to the sea (Table 1). Depth and slope measurements were made in 1895, i.e., 55 years after the formation of the breach. Current velocities were calculated on Chezy-Manning formula: $v = \frac{1}{n} H^{2/3} I^{\frac{1}{2}}$ with H = stream depth, I = water surface slope, n = coefficient of roughness. The latter was provisionally set at a constant and uniform 0.037.

The data of Table 1 have been interpolated on the diagram (Figure 4).

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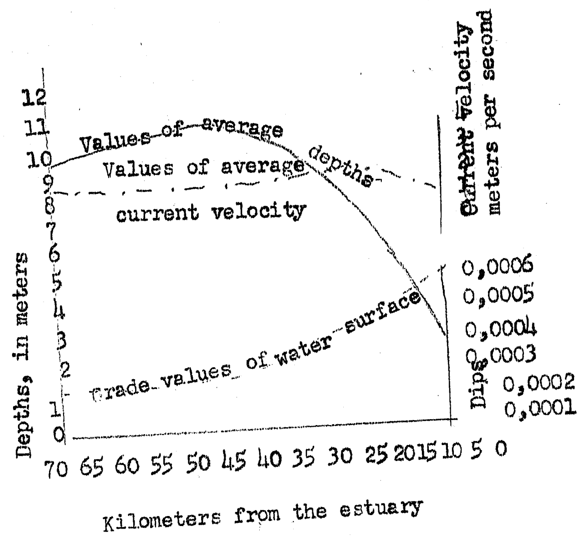

Figure 4. Distribution of average depths, slope values of water surface and average stream velocities in spring floods along the channel in the estuary of one river.

The table and diagram show graphically that the drops in water surface gradually increase downstream, reaching up to 55 centimeters a kilometer at the sea.

Current velocities incr.ase in most areas of spring flood wave spread, but only up to a certain point, below which the velocity subsides due to depth reduction, in spite of the extended increase of slope. The slowing of speed occurs in the sea area where the flow of the stream to the sides and the mingling with sea water produces the formation of a sand bank.

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TABLE I

Distribution of slopes, depths, and current velocities along the channel at estuary river sector in the 1895 spring flood

Distance from the roadstead in kilometers	Slope of water surface	Depth in meters	Stream velocity meters per second
52-70	0,000170	9.8	1.60
41-52	0,000186	10.5	1.68
32-41	0,000191	10.8	1.74
18-32	0,000257	9.9	1.90
9-18	0,000364	8.9	2.20
5-9	0,000457	6.5	2.00
0-5	0,000553	3.0	1.26

SOME EMPIRICAL FUNCTIONS

To determine estuary drop zone length or, as frequently stated, "zones of the spreading of a spring flood wave," we conducted the following research

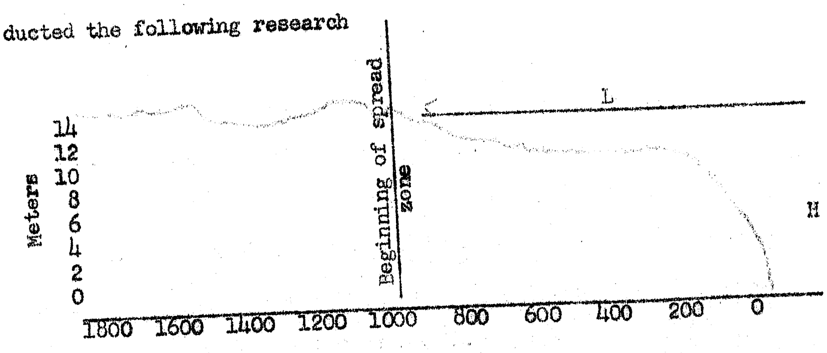

Figure 5. Diagram of the increase of spring flood crest levels over the level of stable mean water level in the lowlands of a river.

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Diagrams of the increase of spring flood crest levels over the level of stable mean water level were made of the lower currents of 15 large lowland rivers of the USSR over a period of several years. Figure 5 is a copy of a similar diagram. As is seen, the uniformity of drop is broken by separate local depressions and crests. Local changes in the estuary drop durve, which depend upon the form of the trough, on the access of water from the tributaries and the formation of ice jams (Ice jams in northern rivers are the main reason for the very pronounced lack of uniformity of the drop curve), sometimes complicate determination of the general curve length. However, in the majority of cases, it was possible to fix the curve limits definitely. The data derived suggest the following equation:

$$L \quad \frac{kH}{h} \qquad (2)$$

with L = length of spread-zone in kilometers; H= level amplitude in meters at the beginning of the spread-zone; h=kilometer drop of the average current of a river in centimeters; k= coefficient, whose value for rivers with slopes of 0,00008-0,00010 is about double, and for rivers with slopes of 0,00003-0,00008 is about 2.5. By increasing spring flood intensity, the coefficient value increased according to the equation $k_1 = k + 0{,}000024Q$, with Q the maximal issue in cubic meters per second (the coefficient equation k=f(Q) was derived for rivers with a spring flood issue from 4 to 60 thousand cubic meters.)

Estuary zone drop lengths in large rivers with small slopes reach several hundred kilometers. Thus it exceeds 2000 kilo-

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meters on the Ob'.

By substituting kilometer drop with the slope in formula (2) course length in meters is derived. Restating the equation as $I = \frac{kH}{L}$ and by substituting $\frac{H}{L} = I_1$, we derive:

$$X_1 = \frac{I}{k} \qquad (3)$$

Presupposing a stream whose bed for the length of the middle and lower current has a uniform bottom slope I, in the area of estuary drop the surface slope will equal $I + I_1$, i.e., 1.5 times greater than in the middle current. Thus it can be approximately calculated that velocity (v_2) of the current in the sector of the drop increases proportionally $\sqrt{1.5}$, i.e, $v_2 = 1,23v_1$, with v_1 - actual current and v_2 -- current in the drop zone. The increase of current velocity toward the river estuary inevitably results in erosion of the bottom which is continued until the depth attains some limiting values. Limiting depths are those in which a given current velocity is unable to continue further erosion of the bottom, but, on the other hand, is adequate for transporting material which has reached the top and for protecting the depth from silting. A frequent occurrence of a similar situation apparently conforms to the function expressed in Kennedy's formula for non-silting beds (It should be noted that in using the formula calculations of Ye. A. Zamarin (1948) the non-silting depth in the estuary is greater by far than that obtained by Kennedy's formula).

In this case, following this formula, we can state the following equation

$$\frac{v_2}{v_1} = \frac{H_2^{0.64}}{H_1^{0.64}} \qquad (4)$$

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with H_1 = actual depth and H_2 = depth in the drop zone.

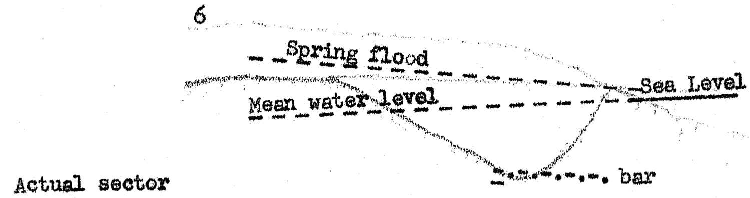
Following the equation (4) and the empirical functions cited above, we can state:

$$H_2 = 1.37\ H_1 \qquad (5)$$

Figure 6 depicts the schema of a longitudinal profile of a river which has an actual depth (i.e., middle current depth) at the spring flood period of roughly 20 meters. The depth in the estuary zone obtained from the equation (5) should be 27 meters.

Actually, estuary zone depths may be considerably greater than those which are obtained from our approximating calculations. In fact, the influence of the estuary drop increases in the degree of approach to the estuary not linearly, used by us for simplification, but in a gradual function, due to which the greatest depths are localized in the lower half of the spread zone.

6



Spread zone

1 ---- 2 ~~~~ 3 -.-. 4

Figure 6. Schema of depth distribution in estuary sectors of a river 1 -- median (rated) levels; 2 -- actual levels 3 -- median (approximate) bottom 4 -- actual bottom

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On the basis of observations, the actual maximal depth of a river (in spring flood) reaches 50 meters. The median depths of estuary depressions equal the doubled depth at the upper reaches of the river, calculated from spring flood level. Depth of isolated estuary depressions sometimes are even greater and in large rivers can exceed 100 meters (the Danube, for example), but these isolated, overdeepened reaches usually are formed by whirlpools resulting from uneven stream broadening. Especially great depths result when the channel is narrowed right up to the estuary, or when there are rapids in front of the estuary; it is then that great rivers with surging spring floods have estuary depression depths reaching several hundred meters (the Congo).

Should the depth in a spring stream at the estuary zone not increase, the depth of the stream, the mean water level of which is known, must nonetheless increase several fold. This is because at the confines of the lower reaches of the estuary zone, the water level at spring flood and the mean water level approximate one another as is seen in the example indicated in Figure 6.

Equation (4) can be transposed, after having revealed the value of current velocity according to the Chezy-Manning formula, yielding:

$$\frac{H_1^{0.64}}{H_2^{0.64}} = \frac{n_2 H_1^{0.67} I_1^{0.5}}{n_1 H_2^{0.67} I_2^{0.5}} \qquad (6)$$

Because the H power indices in both parts of the expression

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are close enough to be equal, the following equation can be written:

$$\frac{I_1^{0.5}}{I_2^{0.5}} = \frac{n_1}{n_2} \qquad (7)$$

Thus, with this depth distribution and current velocities, rendered by equation (4), the slopes are determined by the values of the factors of roughness.

The value n can be replaced (according to Chang) by a mean diameter of a portion of the bed alluvia. Equation (7) will then be expressed:

$$\frac{I_1}{I_2} = \frac{\sqrt[3]{D_1}}{\sqrt[3]{D_2}} \qquad (8)$$

with D_1 the median diameter of bed alluvia in an actual sector, and D_2 the median diameter of bed alluvia at the sector of estuary abatement.

Because every river distributes alluvia along its stream, as a general rule, then, in a river which deepened sufficiently its bed at the estuary, the slope must lessen downstream even during high spring floods. Thus the median diameter of the portions of the bed alluvia decrease in the Volga bed at the stretch of the spread zone almost twofold (comparing the mean roughness of bed alluvia at Kuybyshev and Astrakhan'). Thus for the Volga I_1

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would equal $1,26I_2$. Comparing longitudinal profiles of Volga high waters at the sector of the Kama Confluence -- Saratov and the Stalingrad -- Boasta sector, the ratio approximately conforms to the theory.

Thus at estuaries with a sufficiently deepened bed, the stream slope in spring flood may be somewhat less than under usual conditions. This phenomenon was the basis for the erroneous belief about the purported "supporting" influence of the sea.

The slope at the estuary lessens sharply with a lowered water level (the nature of dip changes depending upon water level fluctuations at the Volga delta were presented in great detail in the monograph of V. Valedinskiy and B. Apollov, 1928. The deepest estuary depressions are formed primarily by water discharges at spring flood peaks. At all other water levels (completely formed in the estuaries), the dips must be less than at the upper (usual) river reaches, because the stream courses along the over-deepened bed, which is terminated by the bar. With a progressive depth increase and decrease in current velocity, proportionate to the proximity to the estuary, the slopes (at low water) lessen to a value sometimes quite insignificant.

Equation (4) substantiates still another function very important in comprehending the morphology of downstream beds. If a river has a uniform water discharge at median and low current (discharge uniformity in rivers at median and low current can often be observed in rivers which pass from more humid to less humid areas), equation (4) is fully correct only given a gradual narrowing of bed downstream. Thus with $Q_1 = Q_2$, $v_1 w_1 = v_2 w_2$; because

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$$v_1 < v_2 \text{ then } w_1 > w_2$$

If cross sections w_1 and w_2 are parabolic, increase of downstream depth is possible only by reducing width. This rule of course cannot be rigid: two such forms of cross section can always be selected that the cross section with less area can have greater depth along the bed axis and a greater width. Nonetheless it is not unusual that at first glance a narrowing of bed and flood valley is observed at the estuary abatement zone.

To our knowledge, V. V. Dokuchaev (1876, page 198) first noted this phenomenon. He contended that in many rivers downstream the number of lake-like river valley widenings perceptibly decrease. Apparently V. V. Dokuchaev implied the downstream area above the delta because the width of river valleys in deltas is considerably greater than in the middle course of a river.

V. M. Lokhtin (1892) determined that the Dnestr middle course had a greater bed width but a lesser depth than downstream. P. T. Tutkovskiy (1893, page 810) noted the very same phenomenon.

According to A. A. Shlyakhtenko's data (1927) the bed on the Lena River from Zhigansk to Bulun gradually narrows more and more in relation to the increase of depth.

The Amazon River flood valley at the middle course has a width of 80 to 100 kilometers whereas at the lower course it is two or three times narrower; thus at Obidos and Santarena (approximately 500 kilometers from the estuary) its width is 30 kilometers.

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"At the eastern area of the basin only a narrow strip of flood valley accompanies the course of the Amazon" (James, 1949, page 349).

The narrowing of the Mississippi's downstream bed is traced with particular clarity and regularity. Data on the bed width of this river will be cited later.

Once again, however, we shall recall that the narrowing of a river bed downstream is not universal; on the contrary in many cases bed width downstream increases. This is observed in estuaries subject to high and ebb tide and wind waves as well as on sectors of descending shores. On the other hand, in rivers whose downstream areas recently left sea level a great widening of flood valley is perceptible -- the traces of the old deltas (for example the widening of the Volga flood valley at Krasnoarmeysk and Vladimirovka).

In all cases, however, even at estuaries, the very singular process of narrowing of the central stream nucleus is observed -- a phenomenon still in want of any satisfactory explanation.

Let us return again to Table 1 and study the column of figures carefully. Towards the upper half of the zone of estuary spring flood spread of a specific river, simultaneous current velocities and depth increases are observable. Consequently, an increase of the specific discharges ensues in the upper part of the curve of abatement in the channel zone. This excess distribution of discharge is possible only by slackening of the current in the marginal areas and the narrowing of the channel nucleus of the stream.

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Several times we noted that, at spring flood in estuary sectors, regardless of the tremendous width of overflow, the channel nucleus of a stream retains the form of a narrow ribbon within which the greater current velocity was noted. This suggests that the very same sort of schema of hydrodynamic structure is applicable in principle for estuary stream broadening which is observed in a diffuser (Figure 7).

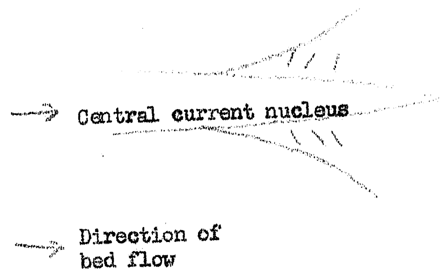

Figure 7. Schema of stream widening in a diffuser

The central nucleus, in similar cone-shaped widening, narrows and occassionally retains great current velocity for a considerable distance. The discharging currents (indicated by arrows) can be observed in the streambed layers and under the influence of these currents alluvia move into the stream's marginal zones where conditions favoring accumulation are created due to slackened velocities of the current. When the zone of channel velocities is narrowed sharply, it begins losing stability: the rapid persistently fluctuates and starts meandering between whirlpools.

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SOME PECULIARITIES OF THE NATURE OF ESTUARIES AND THE NATURE OF OVER-DEEPENED RIVER SECTORS AT MEAN WATER LEVEL

As has already been demonstrated, functions resulting from an analysis of equation (4) are not universal because an estuary type of bed is frequently observed in nature where, in spite of the phenomenon of estuary abatement due to expansion of the cross section, the current velocities gradually diminish towards the estuary.

In all probability, the formation of estuaries is due more to the ebb and flow current actions than to the activity of the river proper. It is also possible that a river, which bears little alluvia with an intensive subsidence of the shore is incapable of filling the narrow bay formed by it, and all the more retains along its estuary quite a marked current, the swiftness of which markedly increases at spring flood when the greater part of alluvia is borne out to sea. The less alluvia borne by a river, the more stable the form of the estuary.

Examples of similar "stable" estuaries are to be seen in the downstreams of rivers flowing into the Ladoga and Onega lakes. The spring floods on these rivers occur much before the lake levels rise, as a consequence of which favorable conditions are created for the spread of the spring flood and the transport of alluvia. They thus retain deep and wide estuaries with poorly developed bars.

V. S. Sovetov and P. A. Kozlovskiy (1927), in surveying a small northern river which flowed into a lake demonstrated that for 20 kilometers of a well-deepened estuary sector of the bed, the stream had a slope which was impossible to measure. Even

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marked summer floods failed to elevate the horizon and increase the slope albeit the current velocity increased appreciably. The horizons in the lower sector of the river increased only with the passing of the peak of the spring floods for a total of but two days.

This phenomenon which contradicts, at first glance, elementary hydraulics (river current velocity increasing without augmentation of slope or increase of stage) is explicable in that the active nucleus of the stream (at lowered river levels) functions solely as the cross section. In these sectors the slope remains constant with an increase in water discharge only until the dead zones in the cross sections disappear. In these zones counter-currents are sometimes observed which have negligible velocity -- at most a few centimeters per second. The force of these counter-currents is insufficient to revolve current meter vanes but quite clearly turns the vane counter to the general current direction in the stream. We observed such weak counter-currents in the estuaries of some northern rivers.

Because a river bottom for a good stretch (for hundreds of kilometers in large rivers) away from the estuary has subsided below the receiving basin, the waters of the latter can infiltrate into the mean water level for a short distance upstream. (the term "upstream" in this instance should be applied conditionally, because in the region of the maximum estuary depths up to the bar, the river bottom has a reverse slope. Thus, the Volga from Astrakhan' to the roadstead for more than 100 kilometers, overcomes a very steep reverse slope of the bottom.)

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Heavy (colder or saltier) water in the receiving basin sometimes infiltrates into estuary depressions in the bottom strata. Artificial increases of the surface layers of currents under water with rough winds are more observable, where during an artificial increase the active river current is constant in the deep stream strata (Valedinskiy and Apollov, 1928). Separate estuary depressions can be filled with whirlpools when mean water level is being taken. Thus, for example, the 36 meter depression depth in the Northern Dvina's downstream stretches is a pool at mean water level with an undertow eddy in its center.

THE DOWNSTREAM FORMATION OF A TROUGH WHOSE DELTAS GRADUALLY CREEP FORWARD INTO THE WATER AREA OF THE RECEIVING BASIN

A sharp decrease of specific discharges in the peripheral sectors of the stream and the presence of a discharging current in the bed strata contribute to intensive accumulation downstream in rivers where a constant accumulation of the river flood valley occurs. B. V. Polyakov (1935) stated that on the Don, Volga, Elbe and Loire, in proportion to approach to the estuary, discharge of alluvia markedly subsides in favor of accumulation in the river flood valleys. This phenomenon is also noted downstream in the Tigris and Euphrates (Wilson, 1925) where estuary canal surveys revealed that only 10 percent of the heavy discharge of a river reaches the bar. It appears that this phenomenon, with few exceptions, must occur in all lowland rivers.

Thus a river which bears alluvia accumulates a flood plain in the estuary spring flood spread zone and gradually forms a chan-

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nel of high water of the very same type characteristic of an average "behavior" sector. The more intensively, furthermore, does the bed sector of the river flood valley accumulate, the median mark of which usually approximates the median mark of the spring flood line (Makkavyev, 1949, page 63.) The banks which build up along the bed more and more hamper the spread of the spring flood stage.

The spring flood spread zone shifts in proportion to the increase in the river flood valley down along the current with first a resultant shift and gouging of deep estuary depressions into the region of the bar, with the upper portion of the latter (facing the river) washed away, and then the point of initial (upper) curve of spread is displaced downstream. When this curve is moved downwards, the depths supported here by the intensified autumn stream begin to decrease and, simultaneously with the displacement downstream of the estuary depressions, an accumulation and elevation of the mean level of the bottom takes place as do the mean markings of the water level in the upper sector. Thus, due to the joint activity of accumulation and erosion, a new sector of river valley forms and the bar moves forward towards the sea.

It is apparent from the schema of Figure 8, that three alternate zones closely linked one with another can be observed in the river bed bearing alluvia and actively building the delta:

(1) The zone of accumulation in the upper half of the region of the spring flood spread zone.

(2) The zone of erosion in the lower portion of the spread region and

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(3) The accumulation zone within the water-receiving basin.

With permanent erosion level these zones gradually move downstream in proportion to increase in delta length.

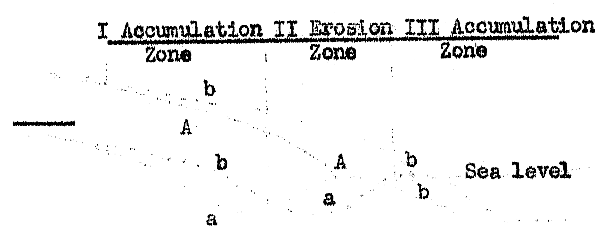

Figure 8. Schema of the re-forming of the longitudinal profile of the estuary sector of a river with a constant erosion level

AA_1, bb_1 -- Successive states of water surface in spring flood in proportion to delta accumulation; aa_1 bb_1 -- successive states of the bottom in proportion to delta accumulation.

The river pushes the bar forward into the receiving basin with the loss of the depths and some elevation of the bottom in the first zone of those enumerated above. The elevation of the bottom will not occur if the delta increases so slowly that the river bed midstream succeeds in dropping to a depth equal to the layer of alluvia which latter should settle because of the displacement of the estuary spring flood spread zone. This, it seems, occurs on the Rhine, Garonne, and possibly on certain other European rivers where an in-

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tensive erosion of the bottom midstream was effected by artificial measures (bed-narrowing by corrective installations) which transformed the longitudinal profile of the river. However, on rivers with an untampered natural cycle, the longitudinal profile within the bounds of midstream can be kept almost constant, because the basic factors by whose interaction the river slope is determined (a drop of locality, the volume and nature of liquid and solid discharge, the form of the valley) -- with a constancy of climatic conditions change very slowly in comparison with the relatively rapid process of delta accumulation. To simplify the following analysis we must conditionally accept the slopes of a river's midstream as constants. It must be noted that this occurrence, of course, will not be usual.

The natural process of downstream river accumulation is accompanied by various phenomena which have complicated it to an unusual degree. The most important of these are the following:

1. River Lengthening, causing the decrease of its general slope and consequently, the transformation of the entire longitudinal profile (more on the processes of transformation of the longitudinal profile at the end of the article).

2. Slope Decrease at River Midstream, accompanied by local increase in spring flood stage. V. M. Lokhtin (1892, page 26) noted the inevitability of this process and stated that because of the wash-out of the longitudinal profile "the high elevations of autumn waters recur all the more frequently."

V. M. Lokhtin's thesis finds some support in an analysis of available data on the maximal spring flood elevations. Thus with

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data on spring flood elevations of rivers of France a tendency towards increasing spring flood elevations downstream in some rivers can be defined. This natural law can be traced with particular clarity for the Loire and the Rhone. Inundations of the Seine for the past decades flood Paris with ever greater frequency (Henry, 1949).

It should be recalled that the highest stage levels on the lower Rhine were observed in the winter of 1940-41 when the main line of defenses of both warring parties were inundated.

Research on spring flood stages for 300 years in South America also indicate that the highest flood levels for most rivers have come in this century (Jarvis, 1939). However, it is impossible to define any tendency towards stage increase of spring floods -- on the basis of these data -- downstream on reaches, and in most of the instances the increase of spring floods are apparantly connected with the lack of uniform discharge resulting from timber felling.

The spring flood level of the Nile at Cairo, on the basis of observations commenced in 622, has a tendency to increase. On the average, it increases from 10 to 15 centimeters a century (Jarvis, 1936). Unfortunately, information is lacking on average level changes of the Nile's upstream reaches.

The increase of spring flood elevations intensifies the phenomenon of estuary abatement and facilitates the increase of relative depths of estuary depressions. Therefore with delta accumulation, regardless of slope decrease, the depth of estuary depressions cannot decrease.

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3. Change of Area of Water Concentration of a river which produces a change in the quantity of discharge, which passes through the lower sector. Sometimes in the process of elongation, two rivers, formerly independent, join in one gulf. The process of a similar junction is noted at the estuary of the Amazon whose delta joined with that of the Rio Para. Frequently, though, on the contrary, the elongation process is accompanied by a loss of water discharge to infiltration, and evaporation; then the elongation of a river does not lead to the increase of basin area because, to the degree of formation of the alluvial plain, separate delta branches are transformed into independent small rivers, encompassing small tributaries of the main river. Examples of such encompassing can be seen in the Mississippi delta (Figure 9).

4. The Splitting of a Bed into Branches. In covering a part of the receiving basin water area with alluvia, a stream concentrates its eroded material along the direction of its channel. The bank which forms initially has an elongated sort of form with a hollow in the upper portion giving it a sandy hillock appearance. Because the discharge of water and alluvia change constantly, situations are always created that the stream, in encountering banks formed at a higher water level, is forced to detour these banks. The spring flood stream, whose dynamic axis in the zones of widening is generally marked by a lack of stability and is able to change its state as a result of any external action, can divert in the future into the detour ravines thus formed. As a result, the lower part of a river can be split into a great number of branches.

In the further advance and gouging into the delta deposits of a riverbed, there is an extinction of a considerable portion of

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the branches; the tendency towards their formation, nonetheless, remains, because in the formation of a valley, as shown above, the river current deposits alluvia primarily in the direct proximity of the channel. A narrow littoral embankment is formed beside the bed which can easily be breached by the spring flood waters, and the elevation of the latter increases all the more with the elevation of the level of the embankment ridges.

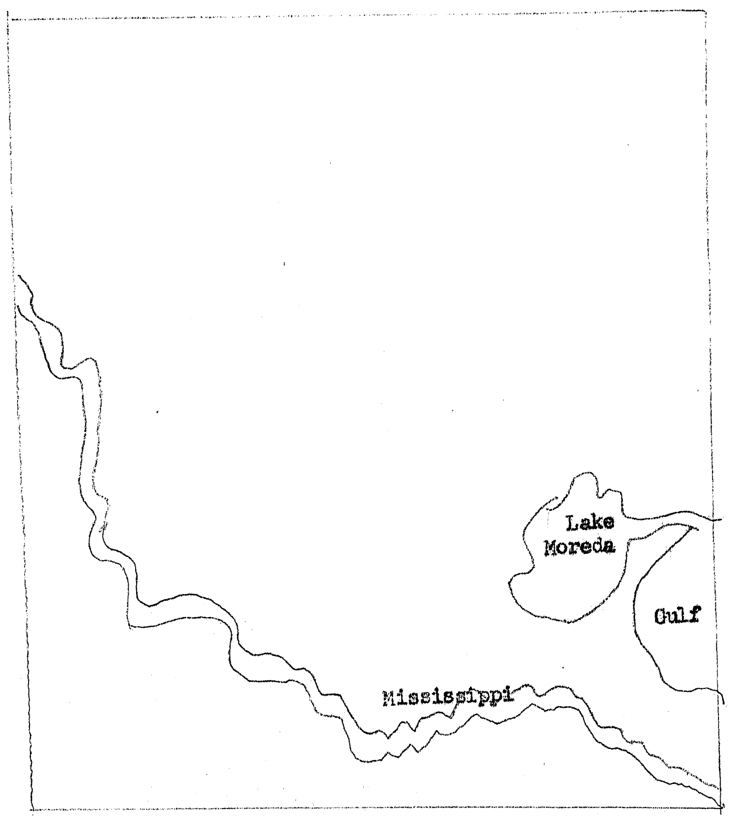

Figure 9. River system in the downstream valley of the Mississippi.

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Embankment breaches and passages of streams from one delta sector into another are natural and characteristic for all actively growing deltas. The length of a lowland sector and the average dip frequently change in accordance with the state of the main current of the river in the delta. As a result of the splitting of a bed into branches the curve of spring flood stage spread becomes very complex and is broken down into a number of separate drops and steeper points.

Without lingering here on other factors (influences of vegetation, wind, the condition of the sea, ice, etc) significantly changing the water in the basic process of the development of a river lowland, the schema of which river is presented in Figure 8, we shall, nonetheless, attempt to clarify by actual example the possibilities of application of this schema for a bed analysis of a natural river.

As an example let us take the lower part of the Mississippi. To the degree of ploughing of soil and felling of timber in the Mississippi basin, the amount of alluvia entering into the river, increase tremendously, after having caused a pronounced delta accumulation (the eroded material of alluvia moreover was augmented by the caving in of the river flood valley and by great dredging projects,which was particularly rapid in the past century when its average forward movement reached 350 meters a year. Man intensified the process of the creation of a river valley and the displacement of the delta, which, under natural conditions, surely would have taken a millennium.

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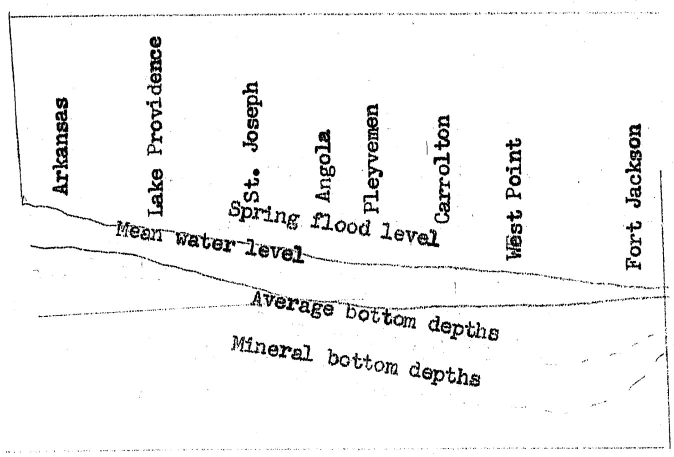

Figure 10. Schema of Mississippi Longitudinal Profile, 1939

TABLE 2

DEPTH SIZE AND PROFILE SLOPES OF THE MISSISSIPPI

River Sector	Distance from Benchwork in Kilometers	Slope	Depth in meters	Current velocity meters per second	Slope	Depth in meters
Arkansas	1060.0	0,00022	25.2	2.96	0.00020	5.6
Lake Providence	892.3	0.00018	27.1	2.76	0.00018	10.3
Saint Joseph	699.7	0.00015	28.3	2.70	0.00012	16.0
Angola	525.8	0.00013	29.6	2.70	0.000027	25.5
Playkvemen	389.5	0.00012	31.2	2.81	0.000013	30.0
Carrolton	213.3	0.000119	32.1	2.93	0.000022	30.6
West Point	126.4	0.00016	26.5	2.92	0.000035	26.2
Fort Jackson	78.1	0.00003	16.2	1.05	-	-
Bar	-0					

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Figure 10 is a schematic profile of the Mississippi from Arkansas City to the sea coast. The depths and water level readings in the spring flood zone and at mean water level have been taken from pilots' charts published by the River Ways Service in 1939.

Current velocities have been determined by allowing for the channel portion of the stream, where it was borne in mind that average roughness of bottom sedimentations for the extent of the sector under consideration changes almost twice, and therefore the coefficient of roughness was made a variable in accordance with Chang's formula

$$n = k\sqrt[6]{D}$$

The following analysis can be drawn from Table 2 and Figure 10:

(a) The greatest average depth both in spring flood and at mean water level is in a sector 126-213 kilometers from the estuary;

(b) The current velocity in the Arkansas City -- Saint Joseph sector noticeably decreases; apparently there is an alluvia accumulation in the channel at this area.

(c) Current swiftnesses in the Saint Joseph -- Pleykvemen sector are practically constant; apparently there is a neutral zone here where the transit of alluvia prevails.

(d) Current velocities at the Pleykvemen -- West Point sector increase; apparently this is a zone of actively growing estuary depressions;

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(e) The velocities in the West Point-- sea coast sector subside; apparently this is the zone of bar formation.

Thus in the lower parts of the Mississippi are noted: the zone of accumulation before the spread region (zone I in Figure 8) and the zone of erosion(zone II) in the downstream parts of the spread area; the process follows the basic scheme indicated in figure 8 (the computation method adopted by us is so approximate that the table, of course, presents only the general nature of the functions, and it is impossible to judge the precise zone boundaries).

The Mississippi bed of known mean water level within the spread zone, in degree to the approach to the estuary, narrows down. Its width is: Cairo 1350 meters, Natchez 1200 meters, Baton Rouge 900 meters, New Orleans 750 meters. This example fully confirms the tendencies towards narrowing of a downstream river bed where the river forms its bed within the alluvial plain.

TABLE 3

CORRELATION OF EXTENT OF ABATEMENT IN SPRING FLOOD AND AT MEAN WATER LEVEL AT VARIOUS SECTORS OF THE LOWER VOLGA

Bench Marks	Inter-bench mark distance in kilometers	Average Abatement Centimeters/kilometer				Correlation of Spring and Mean Water Level Abatement	
		At Flood peak		At Mean Water level			
		1912	1926	1912	1926	1912	1926
Astrakhan-Stalingrad	496	3.78	3.82	2.71	3.07	1.40	1.21
Stalingrad-Kamyshin	179	4.70	5.48	4.00	3.54	1.18	1.54
Kamyshin-Saratov	237	4.06	4.36	3.87	3.94	1.05	1.11

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[The profile dip of water surface at spring flood was determined as the average between the dip at the end of peak and at the beginning of the recession)

The question on requisite criteria arises for the practical determination of the extent of zone spread, in connection with the fact that downstream on rivers with a fixed geomorphological cycle, slope may be even lower at spring flood because of large depressions and a decrease in alluvia coarseness, than in the usual sectors. N.S. Lyelyavskiy (1895) pointed out that the basic criterion for gauging an estuary zone was the correlation of spring flood mean water slopes with those at mean water level. If this correlation is greater than the unit, a preponderance at given points of deep water stretches is indicated, the existence of which are due to estuary erosion. An example of the slope distribution on the lower Volga is indicated in Table 3. The figures show that both in the exceptionally high spring flood in 1926 and in the average flood of 1912, the correlation between the extent of abatement into the high water and into the mean water level is greater than the unit along the entire sector Saratov-Astrakhan.

INFLUENCE OF EROSION LEVEL FLUCTUATIONS ON THE GEOMORPHOLOGICAL CYCLE OF A RIVER DOWNSTREAM

Because the location of a spread zone is linked with the source of its existence, i.e., the estuary, the dislocation of the latter should thus be accompanied by a dislocation of the accumulation and erosion zones for the length of a river. Thus erosion level changes (when accompanied by a change of estuary site) (It is true that

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erosion level will change as a result of tectonic phenomena, but the length of the river remains unchanged. However, where an erosion level change is wrought by level fluctuations in the water receiving basin or by a general change of dry land elevation, the length of a river cannot but change.) are very indicative of the bed cycle.

First let us examine the simplest schema of a phenomenon, having been adopted for the first such instance, when with a change of erosion level, a simultaneous increase of both length and abatement of a river ensue in such correlations that the average slope is unchanged (Figure 11). The lowering of erosion level thus generally results in the very same phenomena (only they are more intensive), as in its static condition: the doredelta is gradually notched, the bar moves forward into the sea and the estuary depressions simultaneously move forward down along the stream. At the very same time there is a filling from the upper side of the old depressions and an acculation of alluvia in the river valley. The whole difference is that the bar will move more intensively and the acculation layer (cross-hatched in the sketch) will be somewhat less. Moreover, the magnitude of an accumulation layer for a given period will be dependent upon the rate of lowering of the water basin level. The magnitude of alluvia deposited in the bed on the average should not exceed the extent of amplitude of the levels, but because the separate estuary depressions attain a very great depth, in places the alluvia layer at various isolated sectors can be significant (up to 100 meters and more).

There are two contingencies in increasing erosion level: (1) abrasion widens the flooded valley so that the spring flood wave spread will be unimpeded at the sector of the new estuary; (2) abra-

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sion does not widen the valley and the estuary acquires the form of a narrow estuary.

In the first instance, upon increasing the erosion level, the process is the exact converse of that in Figure 11; the estuary depression moves upstream; the erosion-formed alluvia fill the upper dip of the bar and the latter likewise moves upstream. Because the spread zone is diverted in a direction counter to the stream, a deepening of the river bottom is possible and even a certain lowering of level.

In the second instance, the spring flood wave spread is concentrated at the expanse of the estuary. We shall not take up this instance because of the very complex nature of the phenomenon, but it seems that Davis's schema is applicable here, particularly if the valley is narrow and canyon-shaped.

[See figure 11 on following page]

In rivers bearing much alluvia and actively building deltas, with a very gradual elevation of the level, the length can remain constant and acculculation will occur over the length of the river lowland. Thus the elevation of erosion level can result in a general accumulation in the bed and river flood valley even when it is very gradual and the rate of delta accumulation is equal to or exceeds the rate of the advance of the shore line of the sea.

The rapid increase of erosion level in a wide valley can result in quite a marked erosion of the bottom (to a depth not in excess of the amplitude of the levels, i.e., up to several meters), instead of accumulation and elevation of river level (according to Davis's schema).

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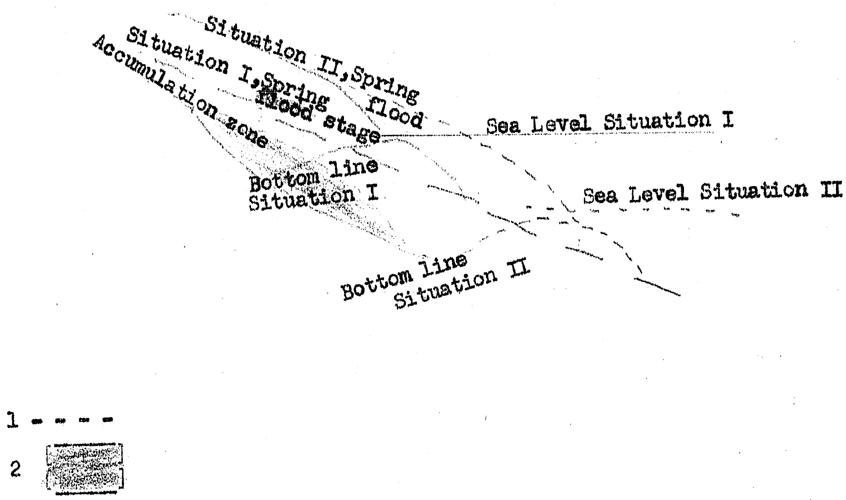

Figure 11. Schema of Dislocation of Estuary Depressions with Changes of Erosion Level

1. General site gradient 2. Accumulation zone with a lowering of sea level

In rivers subjected to sea water inundations in the past, the estuary depressions, apparently, were displaced. Like a plough, they furrowed the bottom of the valley and again filled up with the recession of the basin. In all likelihood, it may be possible to link with the processes of displacement of estuary depressions the circumstances that the great Volga, great Don, great Oka, great Kama and some other "grandparents" of the rivers of the Russian plain had a bottom several meters or several tens of meters lower than the present day-rivers.

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To illustrate the basic schemas (Figure 8, 11) we shall cite two examples of the dislocation of the estuary zone.

1. The Breach of the Vistula Delta (Figure 12b) occurred on the night of 31 January 1840 as a result of an ice jam formation. By the following morning the width of the breach was 300 meters which then rapidly expanded to 750 meters. In the sector above the breach a typical recession curve formed (Figure 12,A). Because of the tremendous erosion of the bottom, the water level of the stream (whose mean water level was known) dropped to 74 centimeters and the depth increased markedly. The deepening then spread several tens of kilometers upstream, after having caused a gradual tapering-off lowering of the stage of mean water level.

The example cited may not be very convincing with respect to a simple explanation of erosion only by the increasing of slope due to a shortening of river length with unchanged sea level. But one should remember that an inundation by the sea of the lower sector of a river with a slanting longitudinal profile will always be accompanied by an increase of its average slopes.

It is exceedingly interesting that in this case the fixing of the abatement curve of the water level was possible to a degree as thorough as would be seldom observed even under laboratory conditions.

2. The Lower Volga. The question of the lowering of the Caspian Sea level has been so well aired in contemporary literature, that we will not halt to describe this phenomenon here. It should be noted that the lowering of the sea resulted in the increase of river length in the past 50 years by 25 to 30 kilometers and in a certain

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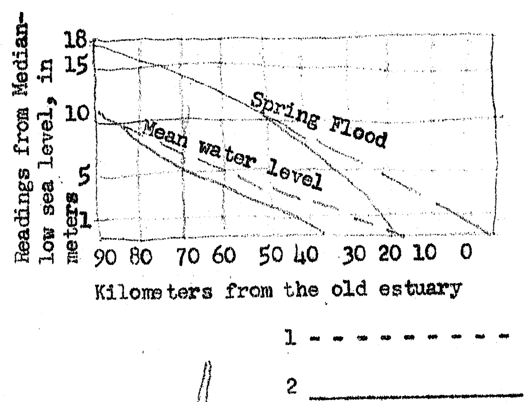
A



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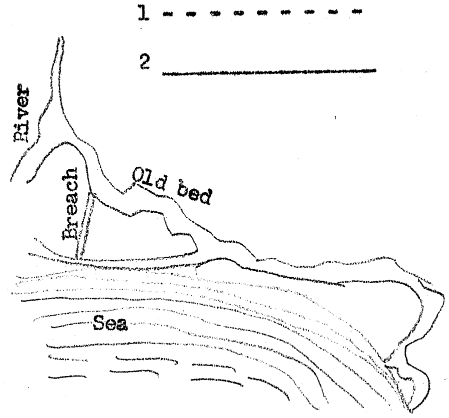

Figure 12. The 1840 breach of the Vistula downstream

A - Longitudinal profile of water surface at spring flood and mean water level before and after the breach; 1-levels before breach; 2- Levels after breach B - Sketch indicating old and new beds.

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increase of slope which slightly increased the average slope of the Volga downstream. Thus, as with the preceding example, with an increase of dips the estuary does not move up, but downstream.

In regard to the shifting of the estuary, at present an intensive gouging of separate beds of the delta branches of the Volga and an advance of depressions to the sea is observed. According to researches conducted at one of the branches of the delta the shift of the deep depressions follows immediately upon the movement of the shore line. Figure 13 depicts a schema (the schema was drawn up by Ye. F. Belevich who obligingly authorized its publication in this article) of the change of a longitudinal profile of the bottom of a branch for the period 1823 through 1947. This branch is second grade and general delta reformations had comparatively little effect upon it. Ye. F. Belevich notes, too, that washing away of the upper parts of islands is very typical for the lower sections of the delta, which, in our view, usually is a morphological attribute of an active, deepened erosion for flatland rivers. The processes of erosion of the bottom resulted in a lowering of the mean water level horizon at Astrakhan', which however cannot be said for spring flood waters.

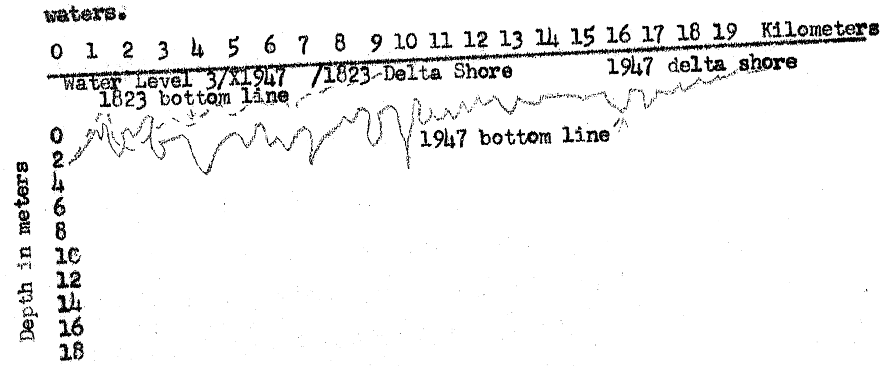

Figure 13. Re-forming of the bottom in one of the branches of the Volga delta, which resulted from the receding of the sea. The 1823 longitudinal profile of the bottom is indicated by the dotted line., the present longitudinal profile by the solid line.

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The gouging of river estuaries in the area of the fore delta with the lowering of the erosion level, in the light of the picture obtained is not something unusual, and another aspect of the bed process linked with the lowering of the erosion level interests us more, to wit: accumulations at the river sector above the estuary depressions (in the first zone, Figure 8). Volga researchers have paid very little attention until the present to this latter phenomenon, apparently not assuming the possibility of a functional connection of accumulation and the raising of a river bottom with the lowering of the erosion level. B. V. Polyakov alone, in one of his last works (1946, page 419) shows that "on the lower Volga a slow raising of the bed was detected, attested to in the raising of annual horizons approximating one centimeter a year."

We attempted by other means to verify the authenticity of the phenomenon noted by B. V. Polyakov.

1. The rated, or a cut, horizon (the provisional horizon to which are pegged the readings of the bed surveys, and conforming to which the depth of dredging in a ship channel in the deepening of sand banks are fixed) is a quite adequate index of the average bottom level. The reading of a rated horizon are set by the method of so-called "instantaneous" leveling for the entire extent of the navigable portion of the river with painstaking care, because errors in computing it result in tremendous non-productive expenditures in dredging. In checking the cut horizon at numerous bench marks of the lower Volga, the Volga River Ways Service succeeds in raising this horizon at 30 bench marks on an average of 50

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to 70 centimeters. Figure 14 shows the magnitude of changes of cuts from 1938 horizon levels. It is apparent that the changes of the cuts are not uniform. Apparently this was caused by local re-formings of the bottom, but almost all were in the nature of an excessive increase.

Excessive increases of cut horizons are established relative to bench marks located along the shore at short distances (usually less than a kilometer) from the shore waterline. Therefore it is impossible to suppose an influence of tectonic causes for this phenomenon, because a similar protrusion of the Volga bed relative to the shore or, on the contrary, the subsidence of the latter in comparison with the bed can occur only where along the channel, strictly conforming to its fold, the axis of the anticlinal elevation would extend.

[See figure 14 on following page]

2. In the area from Kuybyshev to Vladimirovka the silting of river ports and coves is exceedingly great. The volume of alluvia deposited annually per unit of water area is, on the average, 1.5 to 2 times greater than at the midreaches and 2 to 3 times greater than at the upper Volga. Data on silting from computations on dredging and from comparative calculations on water area volumes, were analyzed by us for the period from 1923 through 1945 (Makkaveyev, 1951).

3. The fact of the process of alluvia deposit in certain parts of the Volga downstream is confirmed by direct observations after a heavy river discharge. According to G. I. Shamov's data

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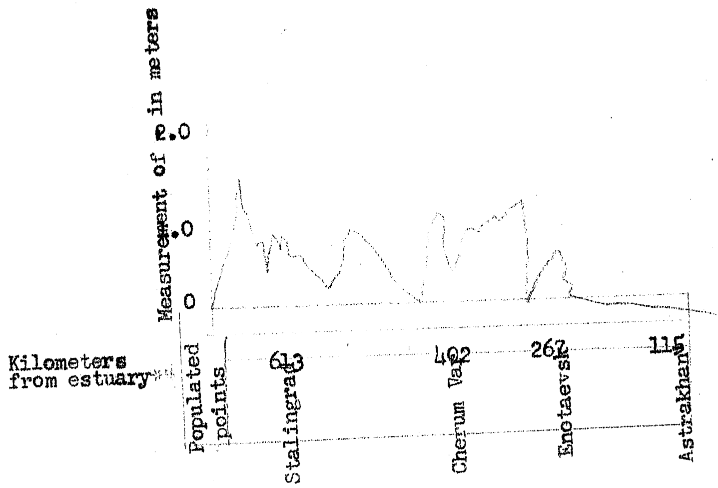

Figure 14. Measurement of cut horizons on the lower Volga relative to the 1938 horizon. Above the line of abscissas are plotted the increase, below -- the decreases of the cut horizon. The cut horizon up to 1938 was taken for zero on the chart.

(1949), the average alluvia discharge at Kamyshina is 950 kilograms a second, whereas at Dubovka it decreases up to 597 kilograms a second. Thus at this sector there is a 353 kilogram accumulation per second, or about 11 million tons annually.

4. Usually the growth of river islands, sand bars, and banks upstream is a graphic morphological symptom of intensive accumulation in the bed. One of our articles (Makkaveyev, 1948) cited comparative sketches of Golodnyy Island near Stalingrad, the above-

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water part of which moved upstream at an approximate rate of 50 meters a year. Regressive sand bank growth was noted at Saratov, at the issue of the Akhtuba, near Vladimirovka and at a number of other places of the lower Volga. We call to mind that below Astrakhan' the inverse phenomenon is observed, i.e., above-water parts of river islands are washed out.

Thus many symptoms show that the bottom of the Lower Volga is raised due to accumulation, in spite of the lowering of sea level.

Because one of the main reasons for the lowering of the Caspian Sea level was the decrease of river discharge, it may be presumed that the elevation of the level of the Volga resulted from the diminution of water discharges and thus by the waning of the river's attraction power. This cause, indubitably, had a telling effect upon the phenomenon of the elevation of the bottom; but if it be primary cause, the elevation of the bottom could be traced for the entire extent of the Volga, Kama and Oka, where diminution of water discharges were observed. Moreover, an analysis of the changes of cut horizons which occurred in the past 30 years, and an analysis of low water level horizons for the last 70 years indicate that others, like these, vary around a certain average state, without revealing a fixed tendency towards lowering or raising. The spring flood levels for the 70 year period disclose the very same picture for the middle and upper Volga, the lower Oka and the lower half of the Kama. A regular spring flood level elevation is traced only on the upper Oka, whereas at the middle Oka the spring flood levels for the very same period reveal a slight

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tendency towards lowering. A progressive and quite powerful (approximately 1 meter a year) lowering of spring flood levels has been established for the upper Kama.

Then too, a comparison of numerous fixed water discharges established for the middle Volga by a survey group in 1882 to 1884 at ~~the~~ village of Rabotka (2296 kilometers from the estuary) and at the town Vasil'surska (2181 kilometers), with current data indicates that a deepening of the bottom by almost 1 meter occurred during the last 50 year period.

Thus the elevation of the lower Volga's bottom is a local phenomenon and is traced for the extent of the upper half of the spring flood spread zone (zone I in Figure 8). Below this zone a deepening of the bottom should occur and this is actually traced on data of the water-gaging station at Astrakhan' and has been ascertained by direct observations at the delta branches.

The phenomena currently observed on the Volga generally confirm the theory developed by us on the mechanism of erosion and accumulation downstream on the Niman, apparently, a raising of the bottom is also occurring. S. I. Kollupaylo (1933), from an analysis of water surveys from 1811 through 1930, concluded that an elevation of the bottom by about 30 centimeters occurred for this period. The process of the recession of the sea in the region of the issue of the Niman is several times less intensive than the process of the recession of the Caspian Sea. Nonetheless at Klaipeda the relative elevation of dry land is estimated approximately equal to 1 millimeter a year (Belousov, 1948, page 110).

The theoretical analysis and examination of examples convinces us that the idea of a tributary of the river stream with the waters

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of the receiving basin has little foundation. However, in a number of cases the gouging of rivers with the lowering of erosion level, indubitably, does occur and, for example, if the Volga and Ural were not gouged in the lowering of the Caspian Sea level, then they would spread in the flatland like the Bol'shoy and Malyy Uzen', Uil and other small rivers of the Caspian Plain. The fact is, apart from processes connected with estuary abatement or tributaries which produce only local and, virtually, not very extensive gouges or silting of the bed, the process of general change of river profile exerts a great influence (and in the case of significant fluctuations-- the decisive influence). Where in the recession of the receiving water basin, territory crops out having a greater slope than the average slope of the water surface of the stream in the river midstream (in the forming of the river discharge), a general gouging of the bed, initially downstream, then midstream ensues. Where the river dip exceeds that of the cropped-out territory, an accumulation of alluvia in the flatlands ensues, as does the obstruction of the estuary and a spreading of the river along the plain. The greater the river water discharge, all things remaining equal, the slighter the water surface slope and the greater, probably, the gouging of the river. Small rivers have smaller slopes and frequently are subject to silting in the lowering of the erosion level only if they do not become, in the changes of the shore line, tributaries of another more powerful river. The change of a river's watershed has great significance here (as noted previously), as do climatological changes which resulted from the changes in erosion level or what accompanies it. All these processes of great complexity are not comprehended within Davis' schema.

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The "fathers" of American geomorphology -- Powell, Gilbert, and Dutton -- conceived of the idea of a close functional link between changes of erosion level and river erosion, apparently because they observed this phenomenon in the region of the Colorado plateau, where in the recent geological past, great vertical upheavals of a block type occurred which resulted in intensive gouging of several rivers. Davis borrowed this idea and without valid reason adopted it as a general principle.

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THE FORMATION OF AN ABATEMENT ZONE AT THE ISSUE OF A RIVER INTO A RESERVOIR, WITH THE ISSUE OF ONE RIVER INTO ANOTHER AND IN SECTORS OF THE WIDENING OF A VALLEY

In cases when the water reservoir fills in the period of the spring flood and its water volume is very great compared with the water discharge into the river at mean water level, conditions arise similar to the case of the issue of a river into a lake. A minor lowering of level at the confluence can occur at mean water level periods resulting from formation of local abatement zones. The abatement phenomenon is particularly distinct when the tolerance capacity of the reservoir exceeds the discharge of water in the river. This is apparently when the thinning-out zones of the tributary, which are actually observed, are located sometimes much below where they have been estimated. The depressions frequently drop lower than the usual ones with the resultant great difficulties encountered in navigation in the upper reaches. The character of such zones is still not the least bit well known, and in researches of rivers at mean water level, an erosion of the bottom can be detected at the confluence area of the stream with the reservoir waters, although, at first sight, an alluvia accumulation ought to occur there. S. V. Rusakov (1948) encountered a similar phenomenon in conducting researches at the upper reaches between sluices of the Dneproges.

The phenomenon of spread of the spring flood has great significance on the character of tributary estuaries and is the

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more explicit the greater the difference in the tributary discharge and the river receiving same. We dealt with the particulars of the geomorphology of river confluence junctions due to this phenomenon in an earlier work (1949) and make but brief mention of them here.

The following have the greatest significance on estuary erosion at river confluence junctions:

(a) ratio of the levels of the amplitude of the confluent rivers;

(b) ratio of the periods of passing of the spring flood peaks.

A clearly marked abatement curve is observed at spring flood in rivers with great level fluctuations which issue into rivers with a less variable level, which furthers the development of the deepened erosion. It is because of this, for example, that the Tosno river which issues into the Neva and which has an insignificant mean water level discharge but a relatively high spring flood so deepened the lower sector of its stream that at mean water level almost up to Nikol'skiy Village it reaches the tributary and has great depths. In instances where the spring flood in the tributary passes much earlier than in the main river, a deepened erosion also usually develops in it downstream. In regions with monsoon type precipitations particularly when large rivers gouge coastal mountain ranges, it is characteristic that tributaries in the lower regions

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of river basins experience a flood earlier than the main river, while in the upper regions of the basins the tributary floods are somewhat delayed. That is why the estuaries of the tributaries in the lower portion of the basin are relatively much deeper. Such a picture of the Amur was the cause for several researchers assuming a dropping of the downstream area of the Amur valley.

22 Aug 1949

100 60 Centimeters

Figure 15. The formation of a depression of the water surface at the sector of local bed widening. The average depth is 1 centimeter; readings of water relief are given in hundredths of a millimeter from a provisional zero

An intensive deepened erosion in the estuaries of the tributaries of the upper Volga started when, as a result of the construction of water reservoirs, the Volga flood, held off until they were filled, began to wane later than usual. Rivers formerly tranquil in spring flood become raging streams and strongly erode the bottom and shores. Even a tributary as powerful as the Oka changed the spring flood cycle, which has a telling effect also in a tremendous change in the channel in recent years, and in a marked regeneration of landslide phenomena on the undercut sectors of the main shores.

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Finally, the spring flood stage similar to that of the estuary, except that it is much the weaker, can always be observed in sectors of local valley widening. This spread causes the formation of a deep stretch of water at the upper narrowed sector, and at the gorge outlet -- sandbanks of peculiar form of the type of a submerged sandy hillock. We call such sandbanks "external bars" (1949). The phemenon of local abatement at the outlet of a stream into a widened state is readily reproduced in a laboratory. Figure 15 depicts a relief of water surface in the widening of a stream in a pan. The depression of water surface is clearly noticeable at the beginning of widening, shifting with the elevation of level before the narrowing.

INFLUENCE OF SPRING FLOOD TRANSFORMATION ON A RIVER'S LONGITUDINAL PROFILE

One of the best river estuary researchers, B. A. Apollov, has indicated repeatedly that it is impossible to sever a river from its estuary. This profoundly valid thesis, however, is ignored by some researchers who forget that a river in its entirety is a surprisingly coordinated hydraulic mechanism sensitively reacting to changes of external conditions but always in such manner that the separate parts of this mechanism are closely linked by definite, albeit constantly changing, relationships. In studying separate parts and separate forms of the bed character, to avoid errors, it is necessary always to describe clearly the arrangement of these individual phenomena in the general system of the river mechanism.

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An example of a methodically inaccurate, purely formal approach to the study of rivers is the work of Iovanovich (1940) on the investigation of a river's longitudinal profile which won wide popularity. After having quite arbitrarily dissected a river into a countless number of fragments, Iovanovich investigates each of these fragments separately, completely uninterested in the place each fragment has in the whole system of the current. A tree can just as successfully be studied which has been hacked into tiny bits and mixed so that it is impossible to determine to which part of the tree they belong: roots, trunk, or pinnacle.

Moreover, the formation of a longitudinal profile of any sector of a river in great measure depends upon whether it is closer to the estuary or the headwaters. Galileo had already established that in the interaction between a stream and its bed the general form of the longitudinal profile is gradually developed and is characterized by a subsidence of slope from the headwaters to the estuary (parabolic curve exposed by the curvature downwards).

There are many causes contributing to this form, the most essential of which are:

(a) the increase of downstream water discharge, as a consequence of which friction losses decrease towards downstream, and the stream's bearing capacity does not diminish regardless of the lessening of slope;

(b) the distribution, abrasion, and dissolving of alluvia particles, causing the gradual pulverizing of the latter toward the

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estuary due to which the roughness of the bed -- in the same direction -- is diminished, and the current velocity required for alluvia bearing is decreased.

Both these reasons, however, do not always suffice to explain the parabolic form of the longitudinal profile. In many rivers water discharges towards the estuary subside considerably (the Ural, the large Middle Asia rivers, the Tigris and Euphrates, et al), the roughness for the extent of the lower current changes in small degree but the longitudinal profile nonetheless maintains the form of a parabola exposed by the curvature downwards. The longitudinal profile of these rivers would have an inverse curvature (curvature upwards) were it not for the spread of spring flood wave described above.

In the spreading process there is a conversion of a tremendous amount of potential energy into kinetic due to which the stream can effect the same erosion and bearing as in the mid-current of the river, but with a slighter bottom slope.

It is obvious, on the other hand, that the process of formation of the spring flood wave is accompanied by a conversion of kinetic energy into potential and contributes to the lessening of the erosion and bearing capacity of the stream in that part of the river where the spring floodcrest increases downstream.

Figure 16 depicts the schema of the transformation of a spring flood wave along a river. From the source down to a certain

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point midstream the crest rises. The loss of kinetic energy in the elevation of the spring flood wave contributes to a slowing of current in this part of the river, i.e., contributes to a certain "tributary" of the stream in the period of passage of water discharges forming the bed. In the estuary zone where a lowering of the spring flood level occurs, the potential energy accumulated upstream is converted into kinetic energy. We provisionally depicted the slope corresponding to the uniform stream flow by a straight line and designated it average slope. The bottom slope in the zone of formation of the spring flood wave is greater than the average slope when, in the zone of diminution of spring flood, the mean water level slope is less than average. Thus the mean profile of river bottom forms a depression in the midstream zone and the curve of the longitudinal profile is generally denoted by the curve exposed by the curvature downwards.

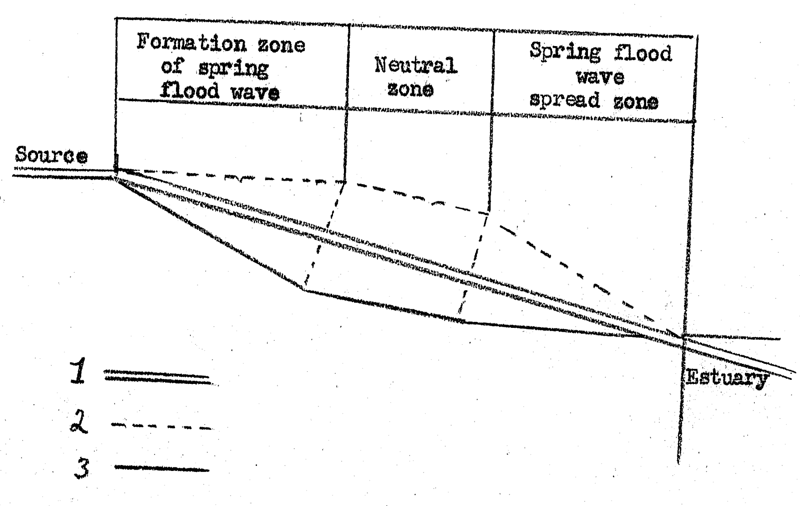

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Figure 16. Influence of the transformation of the spring flood wave on the logitudinal profile of a river

1 -- average slope 2 -- high water profile 3 -- bottom profile

Thus Calileo's hypothesis is accurately and very simply explained.

We can now, oriented by the parabolic profile concept, return to Davis' schema. If with the raising of the erosion level, the basin shore line advances onto the dry ground and a part of the river valley is flooded, then the sloping part of the longitudinal profile is severed. The curve is destroyed and the entire river longitudinal profile must be changed. The sloping sector of the profile must be restored either by delta build-up or by bed-gouging.

To determine which process (gouging or accumulation) will occur, the entire set of conditions must be analyzed which accompany the phenomenon of the raising of the erosion level: the activity of abrasion, swiftness of sea level change, intensity of delta build-up, change of watershed, etc.

In lowering the erosion level, if the lowering is accompanied by a recession of the basin shore or if the average river slope remains constant, the sloping portion of the profile curve should move downwards resulting in a gouging of the bed in the newly formed dry area, but simultaneously there usually occurs an accumulation in

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the midstream junction with the downstream sector. Accumulation and gouging in this case also depends upon many conditions, of which the most important are the change of conditions of discharge and change of general river slope.

The processes of forming rivers' downstream are marked by extraordinary complexity and it may be stated positively that Davis' schema is not applicable in paleogeographic analysis of flatlands (barring their having undergone, in the period in question, great vertical fluctuations). In each separate case a concrete analysis must be made of all factors contributing to the process because, depending upon their configuration, gouging or accumulation of river beds can be indications of totally different processes.

The problem of the character of a stream's longitudinal profile, particularly within estuary sectors, can have a direct bearing upon problems of great importance. The regulation of estuaries and river confluences for navigation, the planning of large reservoirs, the combating of floods, the improvement of river valleys which are subject to inundations -- in all of these burning problems of the national economy, a valid prognosis of the change of the logitudinal profile of a river has decisive significance. All this points to the necessity of a serious and comprehensive study of this problem, to which unfortunately, as yet, insufficient attention has been devoted.

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